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Great expectations: Flow restoration and sediment transport in the Waimea River, Kaua'i

Basil Gomez

Department of Geography, University of Hawai'i, Mānoa, Honolulu, HI 96822, USA

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ABSTRACT

Conventional and novel observations made in the Waimea River basin between 1960 and 1995 permit the total riverine mass flux to be estimated and the influence that flow restoration will have on sediment dynamics in the river's lower reaches to be assessed. Flows between the threshold for sediment transport (~6.0 $m^3 s^{-1}$) and the most effective flow (80.7 $m^3 s^{-1}$) recur annually and transport ~60% of the Waimea River's suspended sediment load. Discharges of this magnitude essentially were unaffected by plantation era agricultural diversions of 2.3 ± 0.7 $m^3 s^{-1}$. The modern-day mass flux from the Waimea River basin is 155 ± 38 t $km^{-2} y^{-1}$, and comparison with an independent cosmogenic nuclide-based estimate implies that it has remained at about this level for the past 10 ky. Previous work indicated that: (*i*) most of the sand the Waimea River transports to the coast is derived from steep, rapidly eroding, sparsely vegetated, bedrock-dominated hillslopes; and (*ii*) the sediment transport regime of the Waimea River is supply-limited at very high discharges (recurrence interval > 2.5 years). Consequently, major floods tend to remove sand from the estuary. Climate change has caused a statewide decline in heavy rainfall, and a commensurate decline in the magnitude of peak flows in the basin's pristine, undiverted headwaters over the past 97 years. The effect this secular change in climate presently is having on streamflow was foreshadowed in the late 1970s by a naturally occurring, warm Pacific Decadal Oscillation phase reduction in the magnitude of flows with low exceedance probabilities. Additionally, the controlling base level at the river mouth has risen and been displaced seaward. Simple proportionality approximations show that, for a constant sediment supply, aggradation will occur if either the magnitude of flows with a low exceedance probability declines and/or base level rises. Thus, anthropogenic stresses on Waimea River's lower reaches are not derived from the within-basin influence agricultural diversions exert on the flow regime and will not be resolved by restoring flow to the river. These stresses primarily accrue from extrinsic factors that will continue to influence the river's hydrologic and sediment transport regimes until global, offsetting, climate-ameliorating measures are implemented.

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1. Introduction

Hawai'i has a long pre- and post-European contact history of diverting river water for agricultural and other off-stream uses (Allen, 1991; Wilcox, 1996). The post-European contact diversions often take the form of screened intakes behind weirs or small, low-head dams constructed on headwater streams. These structures allow flood flows to pass over them but reduce or eliminate base flows in channels farther downstream. This not only reduces the habitat available to the amphidromous native stream fauna, but also places limitations on appurtenant as well as traditional and customary native Hawaiian rights (Oki et al., 2010; Morales, 2016). Consequently, with the demise of monocrop plantation agriculture and the transition to growing diversified, less water-intensive crops in recent decades, proactive community groups have sought to restore water to streams throughout the state (Miike, 2008; Sproat, 2011). Most recently, the Pō'ai Wai Ola hui successfully petitioned for flow to be returned to the Waimea River in western Kaua'i.

Pō'ai Wai Ola hui's petition contended that: 'weak river flows continue and cause more and more silt to fill' the ~2-km-long Waimea River estuary and the adjoining upstream reach; and the sediment cover is 'the extended result of persistently low flows that are insufficient to flush the river bed' (Pō'ai Wai Ola, 2013). The petitioners specifically were concerned about the cumulative impact silt deposition may have on the Waimea River Flood Control Project, which protects >600 homes and commercial buildings, and the lotic ecosystem's continued ability to support traditional cultural and recreational activities throughout the river's lower reaches.

River restoration is a social process that, in the absence of rigorous conceptual foundations and well-tested principles, calls upon science to varying degrees (Wohl et al., 2005). Nonetheless, science provides a basis for informing decisions and emending perceptions of environmental degradation that commonly are underpinned by normative expectations of how river morphology should be, now and in the future. This is because culturally preferred landscape ideals are unlikely to be attained if they cannot be sustained by channel processes (Wohl et al., 2005).

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Low-head diversions have the potential to impact flow conveyance and change in-stream sedimentation processes, but because detailed hydrologic records of diversion operations usually are not available, previous studies often have relied on ergodic reasoning to link a changed sediment response to flow diversion (Bohn and King, 2000; Baker et al., 2011). However, concurrent gauging records were maintained on the Waimea River and the diversion ditches. These records, a profusion of novel supporting information, and several fundamental fluvial geomorphic precepts are amalgamated in an attempt to determine whether or not the "sediment problem" in the estuary and the adjoining upstream reach will be redressed by restoring flow to the Waimea River.

2. Waimea River basin

Kaua'i was constructed by shield and post-shield volcanism, between about 5.5 and 4 Ma (Sherrod et al., 2007). The 221-km^2 Waimea River drains the southwestern sector of the former volcanic center, where basaltic lava flows ponded within the fault-bounded caldera (Fig. 1; Macdonald et al., 1960; Sherrod et al., 2007). Dissected remnants of the caldera's surface form a gently sloping, low-relief plateau that lies at 1200 m elevation and extends northwest from 1569 m high Mount Wai'ale'ale. Mantled by a near pristine montane wet, 'ōhi'a-dominated forest, the ~50-km^2 plateau is known as the Alaka'i Swamp (van't Woudt and Nelson, 1963; Englund et al., 1998). About 4 Ma, dropdown between two major faults on the southern edge of the Kaua'i shield created the 6.5-km-wide Makaweli graben that cuts across the consequent, radial drainage on the shield surface (Macdonald et al., 1960). This concentrated and changed the direction of drainage throughout the ~18-km reach of the river that now flows across the floor of the ~500-m-deep Waimea Canyon to the confluence with the Makaweli River, which maintained a course consequent to the southwestern slope of the graben fill. No perennial tributaries drain the canyon's west wall, and

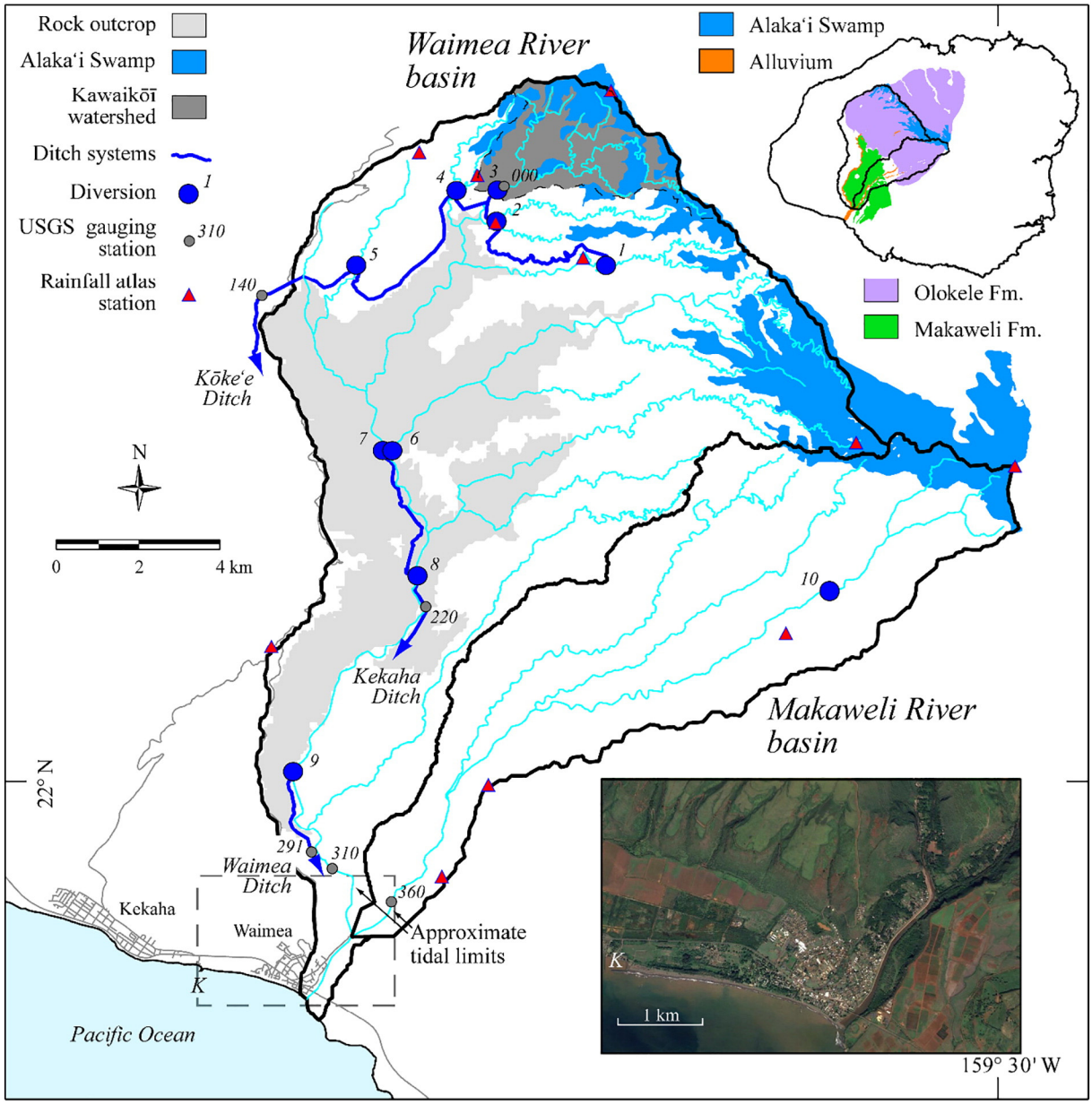

Fig. 1. Major streams and agricultural diversions in the Waimea and Makaweli river basins (distribution of rock outcrops after Hue et al., 2006). Diversions: (1) Mōhihi; (2) Waiakoali; (3) Kawaikōī; (4) Kauauikinanā; (5) Kōke'e; (6) Koai'e; (7) Waiahulu; (8) Waimea upper; (9) Waimea lower; (10) Olokele. USGS gauges: 000, 16010000 Kawaikōī Stream; 140, 16014000 Kōke'e Ditch; 220, 16022000 Kekaha Ditch; 291, 16029100 Waimea Ditch; 310, 16031000 Waimea River; 360, 16036000 Makaweli River. 'K' indicates the location of the Kīkīaola Small Boat Harbor, and dashed box location of inset image. Basin-average rainfall (Fig. 5) was computed using records for all 11 Rainfall Atlas of Hawai'i rainfall stations shown. Insets: top, generalized geology of the Waimea River basin (after Sherrod et al., 2007); bottom, Waimea River estuary and adjoining upstream reach. Google, DigitalGlobe, 2017.

lower gradient hillslopes within the canyon are covered by koa lowland mesic or alien forests. Except for small recreational residence lease lots in the headwaters, and homes and kalo lo'i (taro fields) on the floodplain, land in the Waimea River basin has a forest, wilderness, conservation, or reserve area designation (DLNR, 2014).

On Kaua'i, the interaction between the prevailing Trade Wind flow and the island's topography gives rise to one of Earth's steepest rainfall gradients; and over a distance of 25 km average annual rainfall declines from ~9500 mm at the summit of Mount Wai'ale'ale to ~500 mm on the Kekaha–Mānā Plain. All the major tributaries of the Waimea and Makaweli rivers rise in the high-rainfall Alaka'i Swamp (Fig. 1), where saturated surface soils and low-permeability subsoils encourage rapid runoff (van't Woudt and Nelson, 1963). These factors, in combination with the lack of channel storage and steep basin and stream slopes, produces flashy flood hydrographs that exhibit sharp rising and steep recession limbs and are of a few hours duration (USACE, 1980). November through April are the wettest months (Frazier et al., 2012), and during these wet "winter" months cold fronts associated with upper-level disturbances and abundant tropical moisture also generate heavy rainstorms. The rainfall data utilized herein were downloaded from the Rainfall Atlas of Hawai'i website (http://rainfall.geography.hawaii.edu/) and the National Water Information System (NWIS) website (https://waterdata.usgs.gov/hi/nwis/). Water discharge and discrete sample data also were accessed via the NWIS website.

The Trade Wind flow and rainfall are modulated by the long-period (20–30 year) Pacific Decadal Oscillation (PDO) and shorter period (2–7 year) El Niño Southern Oscillation (Chu and Chen, 2005; Frazier et al., 2012). Kaua'i rainfall is negatively correlated with the PDO and there is less rainfall during the warm PDO phase, which also suppresses 'Kona' storm activity (Frazier et al., 2012). Winters are dry during strong El Niño years and wet during strong La Niña years, and the (dry) El Niño and (wet) winter response to La Niña are enhanced by interaction with the warm PDO phase (Frazier et al., 2012). However, in recent decades, winter rainfall across the Hawaiian Islands has been lower than would be expected from the historical relation with the PDO (Diaz and Giambelluca, 2012), and a decline in heavy rainfall has been linked to weaker Trade Winds and expansion of the Hadley circulation (Kruk and Levinson, 2008; Seidel et al., 2008; Chu et al., 2010; Elison Timm et al., 2011; Garza et al., 2012). Consistent with trends across the state (Oki, 2004; Giambelluca et al., 2011; Bassiouni and Oki, 2013), scrutiny of long-term records shows a significant decline in rainfall and streamflow, as well as in the magnitude of peak flows in the pristine, undiverted headwaters of the Waimea River basin over the past 97 years (Fig. 2). The downward trend in streamflow may also be indicative of a weaker rainfall association with the PDO index and, as evinced by the persistent decrease in the cumulative rescaled departure over the last three decades (Fig. 2D), cold PDO phases in the twenty-first century are associated with extended periods of below-average flow.

During the first quarter of the twentieth century, three agricultural ditch systems were constructed to convey water from the Waimea River and its headwaters to land on and hillslopes above the Kekaha–Mānā Plain leased or owned by the Kekaha and Waimea sugar companies (Fig. 1; Wilcox, 1996). Completed in 1903, Waimea Ditch diverted water from the lower reaches of the Waimea River, at 68.6 m elevation, until it was abandoned in 1990. Constructed in 1906 and 1907, Kekaha Ditch diverts water from the middle reaches of the river, at 243.8 m elevation on Koai'e and Waiahulu streams and 158.5 m elevation on the Waimea River. Begun in 1923 and completed in 1927, Kōke'e Ditch originally diverted water from Mōhihi, Waiakoali, Kawaikōī, Kauauikinanā, and Kōke'e streams above 1051.6 m elevation; however, the Mōhihi intake was abandoned in the late 1980s. Additionally, completed in 1904, Olokele Ditch diverts flow from Olokele Stream (Wilcox, 1996), a tributary of the Makaweli River, at 441.1 m elevation, and small amounts of water also are diverted to irrigate kalo lo'i.

The U.S. Geological Survey (USGS) began to monitor streamflow in the Waimea River basin in 1909, but because long-term records are not required to quantify water-supply potential for irrigation, many gauges operated for only a short period of time (Territorial Planning Board, 1939). Flows in many ditch systems were also monitored from the time of their construction, and to better define current conditions in and flux of water through the entire hydrologic system the gauges on the lower Waimea and Makaweli rivers and downstream from the confluence with the Makaweli River were reinstated in 1943 and

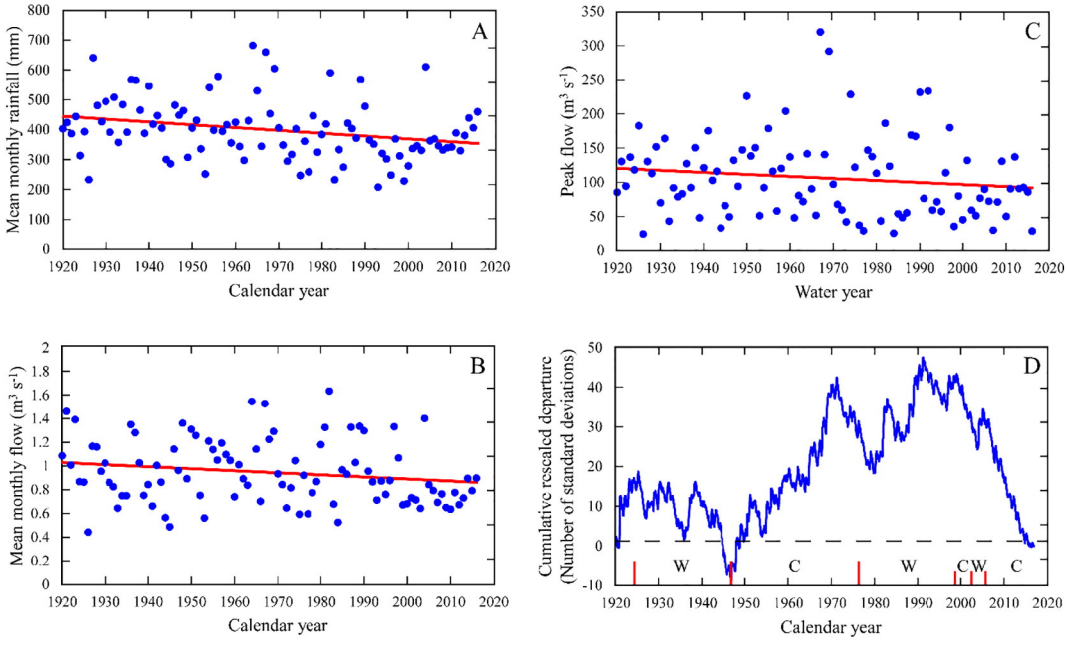

Fig. 2. (A) Trend (significant at the 0.01 level as determined by the Mann-Kendall test) in mean monthly rainfall at the head of Kawaikōī Stream (Kilohana rain gauge), 1920–2016. (B) Trend (significant at the 0.05 level) in mean monthly streamflow in Kawaikōī Stream (USGS 16010000), 1920–2016. (C) Trend (significant at the 0.1 level) in peak flow in Kawaikōī Stream. (D): Cumulative departures of monthly mean flows from the mean of all monthly mean flows (1920–2016) computed and rescaled by the standard deviation of the monthly mean flows. Vertical red lines delimit warm (W) and cold (C) PDO phases. Watershed and gauge locations are shown in Fig. 1.

1944 respectively. The record for the Makaweli River continues to the present, but the gauge on the Waimea River was shut down in 1997 and only recently (summer 2016) recommissioned. Concurrent daily records for these gauges and all three ditch systems exist for the period March 1960 through September 1968 (Fig. 1). Measurements of suspended sediment and total dissolved solid concentrations also were made on the lower Waimea River between June 1975 and August 1995.

For most of its course the Waimea River flows over bedrock mantled by boulders and gravel. The bed changes to sand ($D_{50} = 0.3$–1.6 mm; Copeland, 1990) ~1 km upstream from the confluence with the Makaweli River, and borehole logs show the subsurface material downstream from this point consists of layered alluvial and marine silts, sands, and gravels of undetermined (>15 m) thickness (Fig. 1; USACE, 1981). In the late 1970s the upstream water diversions anecdotally were linked to aggradation within the tidal reach (USACE, 1981), which was widened and partially embanked in the early 1950s. Subsequent flood-control improvements, completed in 1984, raised the existing embankment and extended it to the ocean. Surveys conducted in 1979 and 1987 confirmed that sediment was accumulating in the channelized tidal reach, where the normal and extreme tidal ranges are ~0.5 and ~1.2 m respectively. However, numerical modeling showed that flows in excess of ~450 $m^3\ s^{-1}$ were capable of initiating significant degradation (Copeland, 1990). An emergent, vegetated bar downstream from the confluence with the Makaweli River also was found to be composed of finer sediment than the material present on the adjacent river bed ($D_{50} = 0.2$–0.1 mm; Copeland, 1990). Once mobilized, particles <0.2 mm in diameter typically go directly into suspension (Sundborg, 1956). Thus, the bar's composition is indicative of suspended load deposition. Such bank-attached bars, which form through deposition, augmented by the sand-trapping ability of vegetation, in the region of flow separation / expansion at the downstream junction corner, are a common morphologic feature in post-confluence channels, and the process of their formation is separate from the process of riverbed aggradation (Copeland, 1990; Best and Rhoads, 2008).

The Waimea River transports sand and smaller-sized material to the ocean and the shallow, gently sloping sand reservoir off the river mouth is the largest shorefront accumulation of terrestrial sediment in the Hawaiian Islands (Chamberlain, 1968; Blay, 1999). A small amount of this terrigenous material forms a river mouth bar that is built up by wave action and periodically dissected by the river during freshets or cut through mechanically. Most of the sediment is transported westward until its longshore movement is blocked by the seaward-protruding Kīkīaola Small Boat Harbor breakwater (Fig. 1). Construction of the harbor in 1959 reversed a decades long trend of severe erosion of the ~2.5-km-long Waimea beach, where accretion (1.4 ± 0.7 m y^{-1}) over the intervening 58 years has increased the sand volume by >350% (USACE, 1961; Fletcher et al., 2012; Podoski, 2013; Blay, 2017).

3. Diversions and streamflow

Kōke'e, Kekaha, and Waimea ditches had a collective capacity of 4.6 $m^3\ s^{-1}$. Their combined average daily mean flow for the period of record (March 1960–September 1968) was 2.3 ± 0.7 $m^3\ s^{-1}$, but their capacity was attained on only three days. The effect these diversions had on streamflow is apparent when flow duration curves are used to compare the characteristics of measured flows in the Waimea River to those of reconstructed flows that have been adjusted to account for the effects of the upstream agricultural diversions (Fig. 3). The sum of the mean daily flows recorded in each ditch was added to the measured flow of the Waimea River to obtain the reconstructed flow, and the abscissa is the exceedance probability: $p = r/(N + 1) \times 100$; where p is the percentage of time that a given flow is equaled or exceeded; r is the rank assigned to each flow value; and N is the total number of days in the period of record. At high flows (<10% exceedance probability) there is little or no difference between the measured and

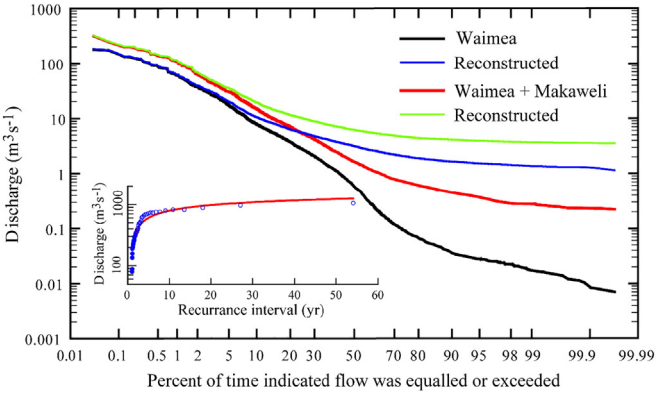

Fig. 3. Flow duration curves for the Waimea River at USGS 16031000 and the Waimea River estuary (Waimea + Makaweli), compiled from mean daily flows for the period March 1960–September 1968. See text for an explanation of the procedures used to derive the reconstructed curves. Gauge locations are shown in Fig. 1. Inset: Recurrence interval of maximum discharge measured at USGS 16031000, water years 1944–1996 years; solid circles denote flows with a ≤ 2 y recurrence interval; and regression line $R^2 = 0.94$.

reconstructed flow of the Waimea River. Based on the silicon content of ground and surface waters, it is estimated that ~30% of the water volume in the lower Waimea River is derived from groundwater (Schopka and Derry, 2012), and the gentle slope of the remaining portion of the reconstructed curve is indicative of the contributions made by direct runoff from persistent rainfall in the headwaters and by groundwater discharge via springs or hyporheic flow through the bed (Macdonald et al., 1960). The surface water diversions have most impact on these low and intermediate flows (>10% exceedance probability). This is also the case in the estuary downstream from the confluence with the Makaweli River (Fig. 3). However, this reconstructed curve is less reliable because it incorporates an estimate of the average amount of water diverted into Olokele Ditch (~2 $m^3\ s^{-1}$; Matsuoka, 1981) and does not account for water used to irrigate kalo lo'i.

4. Suspended sediment discharge

An influential fluvial geomorphic principle is that large floods transport only a minor proportion of the annual suspended sediment load, and that most suspended sediment is transported by frequent events of moderate intensity; where 'frequent' is defined as an event that recurs at least once each year or two (Wolman and Miller, 1960). The cumulative sediment load transported by a given discharge is the product of transport rate and frequency, and the relation between discharge and cumulative sediment load attains a maximum at a high, but not extreme discharge; termed the effective discharge (Wolman and Miller, 1960). Applicable to river basins where sources with high sediment delivery ratios, such as steep, sparsely vegetated hillslopes, gullies and gulches predominate (Trustrum et al., 1999; Hicks et al., 2000), the concept that frequent events drive sediment transport also may be used as an analytical tool to help guide the process of river restoration (Shields et al., 2003).

Using the contiguous flow record for the ~20-year period (June 1975–August 1995) when suspended sediment gaugings were made in the lower Waimea River, a correlation between suspended sediment concentration and water discharge was derived using the sediment rating–flow duration method (Livesey, 1975). Rating curves constructed from concurrent measurements of suspended sediment concentration (C) and water discharge (Q), usually are defined as a power function: $C = {}_aQ^b$; where a and b are empirical coefficients, which can be adversely affected if a large number of concentration values are associated with low and intermediate (nonflood) flows (Gray and Simões, 2008). In the present case only 7 of 160 measurements were made during high flows of <10% exceedance probability, and to alleviate

scatter around the regression line, group averages of the paired independent and dependent variables were computed for the following incremental ranges of water discharge: 1–2, 2–3… 9–10; 10–20, 20–30… 90–100; 100–200; and 200–300 cfs. Suspended sediment-discharge values derived from the resulting regression relation (Fig. 4A; $R^2 = 0.65$) were then determined for the corresponding increments of discharge from the duration curve of mean daily flows. The frequency distribution of daily average values was obtained by multiplying the suspended-sediment load and discharge increments by the time-percentage interval (Fig. 4B). Computed in this manner, the suspended sediment discharge of the Waimea River amounts to 30.0 ± 9.2 t/d, where the uncertainty is approximated by the standard error of the load computed from the year-by-year flow records. Consistent with Wolman and Miller's (1960) principle, ~60% of this sediment is transported by flows between the threshold for sediment transport (~6.0 $m^3 s^{-1}$; Copeland, 1990) and the most effective flow (80.7 $m^3 s^{-1}$), which is 27 times the mean flow of 3.0 $m^3 s^{-1}$, and 200 times the Q_{50} flow of 0.4 $m^3 s^{-1}$. Reference to the maximum series flood frequency curve for the period of record encompassing water years 1944 through 1996 shows that flows of this magnitude recur annually (Fig. 3).

The representativeness of sediment discharges computed from transport curves depends on factors such as the number of concentration-discharge measurements made, and whether or not these data correctly characterize a river's hydrologic and sediment transport regimes (Gray and Simões, 2008). As the level of uncertainty attached to the estimate of suspended sediment discharge suggests, mean flow and peak flow in the Waimea River over the 20-year period in question was quite variable (3.0 ± 1.9 and 367 ± 258 $m^3 s^{-1}$ respectively). Although mean flow is influenced by the diversions (Fig. 3), the influence of rainfall on mean and peak flows predominates, and higher rainfall results in higher streamflow (Fig. 5). Rainfall also drives erosion which, like runoff, increases with topographic gradient, and most sediment is thought to be derived from steeply sloping (30 to 70% gradient), sparsely vegetated, bedrock-dominated hillslopes, broken by intermittent gullies and gulches cut into saprolite and rock that border the deeply incised Waimea Canyon (Gayer et al., 2008). Soils in these locations are thin or nonexistent (Fig. 1), suggesting particle residence times are short. Nonetheless, because chemical weathering may be capable of breaking down most coarse particles, such hillslopes primarily are presumed to deliver sand and finer materials to channels (cf. Sklar et al., 2017). The low (0.1% by weight; Peterse and Eglinton, 2017) amount of total organic carbon in the river sediment also is suggestive of its derivation from these weathering-limited sources.

Larger rainstorms intensify the processes responsible for delivering sediment to channels, and the largest discharge recorded during the 20-year period was 873 $m^3 s^{-1}$; a discharge that was only exceeded by two other floods during the entire 52-year period of record. However, storm activity likely was suppressed by the warm phase of the PDO, which prevailed from 1977 through 1998, and this was the only large flood (recurrence interval > 5 years) recorded in a period that was dominated by smaller events (recurrence interval < 1.5 years). Eight of the ten largest floods on record occurred during the preceding

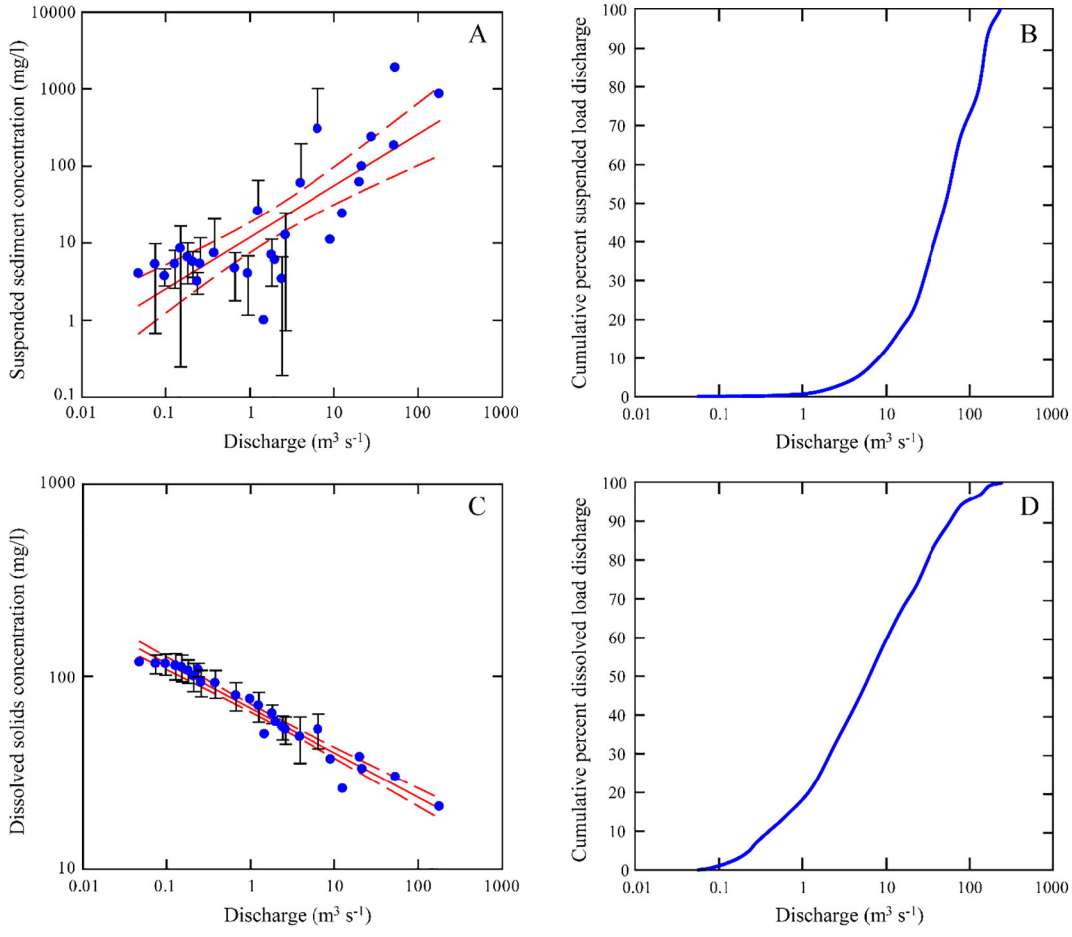

Fig. 4. (A) Suspended sediment–water discharge relation for the Waimea River at USGS 16031000; $R^2 = 0.65$. See text for an explanation of the averaging procedure. Error bars show the standard deviation of the concentration estimate, absent lower error bars denote unbounded lower uncertainties, no error bars indicate a single measurement, and dashed lines delimit the 95% confidence intervals. (B) Cumulative distribution of the suspended sediment load transported at discharges less than the indicated value. (C) Dissolved solids–water discharge relation for the Waimea River at USGS 16031000; $R^2 = 0.95$. See text for an explanation of the averaging procedure. Error bars show the standard deviation of the concentration estimate, no error bars indicate a single measurement, and dashed lines delimit the 95% confidence intervals. (D) Cumulative distribution of the dissolved solids load transported at discharges less than the indicated value. Gauge location is shown in Fig. 1.

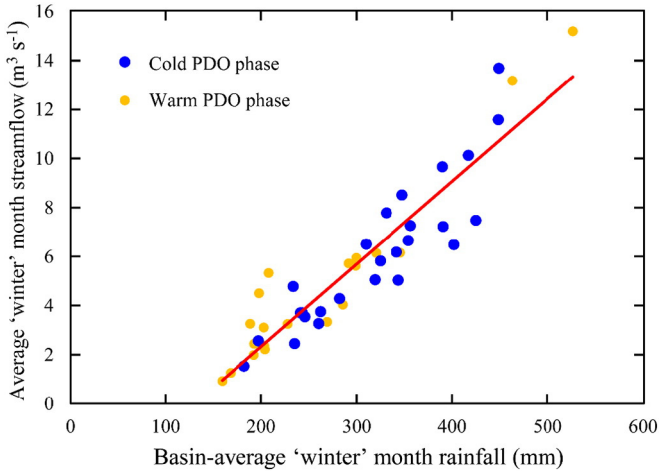

Fig. 5. Relation between basin-average "winter" (NDJFMA) month rainfall and average 'winter' month streamflow at the Waimea River gauge (USGS 16031000), 1947–1996; $R^2 = 0.86$. Gauge locations are shown in Fig. 1.

(1947–1976) cold phase. In general, higher flows convey more sediment, but the sediment transport regime of the Waimea River appears to be supply-limited at very high discharges (recurrence interval > 2.5 years; Copeland, 1990). This is consistent with the pattern of erosion on high-gradient hillslopes, where sediment availability is dependent on the rate at which bedrock loses its structural integrity and is broken down by weathering, such that the rate at which sediment is delivered to streams is limited by the rate of sediment production.

5. Basin-averaged erosion rates

Estimates of long-term erosion rates in the Waimea River basin have been made based on: (*i*) the volume of material removed between the time the pre-erosional topography was constructed and the present (Ferrier et al., 2013); and (*ii*) cosmogenic ^{3}He concentrations in the olivine-rich Waimea River sand (Gayer et al., 2008). Erosion rates inferred from ^3He_c concentrations are averaged over the characteristic timescale of ^3He_c accumulation (which is 10,000 years for an erosion rate of 160 t km^{-2} y^{-1}) and are representative of mass losses over a characteristic attenuation length of 0.53 m for basalt (Cerling and Craig, 1994). Thus, ^3He_c concentrations measured in the Waimea River basin reflect millennial-scale mass loss rates to a depth of a meter or so below the ground surface, that are unaffected by mass losses that may be occurring at greater depth due to subsurface weathering by groundwater (Schopka and Derry, 2012). As such, they reflect the sum of all physical and chemical erosion fluxes that occur in the near-surface environment on hillslopes. However, to compare long-term and contemporary erosion rates, estimates of the bedload and solute fluxes also are required.

Measurements were made of the total dissolved solid concentration between June 1975 and August 1995. Estimated in the same manner as the suspended sediment discharge, using group averages of the paired independent and dependent variables computed for the aforementioned incremental ranges of water discharge and the contiguous flow record, the solute discharge of the Waimea River amounts to 10.1 ± 5.0 t/d. Although the concentration–discharge relation is extraordinarily coherent ($R^2 = 0.95$; Fig. 4C), the greater uncertainty associated with this estimate accrues from the relatively large (~35%; Fig. 4D) portion of the dissolved load that is transported by flows less than or equal to the mean flow (3.0 ± 1.9 m^3 s^{-1}) when the runoff dilution of solutes in base flow generated from groundwater is least pronounced. This is also the flow range that is most profoundly impacted by agricultural diversions (Fig. 3).

Measurements of bedload discharge are difficult to make, and bedload rarely has been monitored directly on a continuous basis (Hicks and Gomez, 2016). Nonetheless, information of comparable quality to that yielded by direct measurements may be obtained by monitoring changes in the volume of sediment stored in morphological units, and repeat topographic surveys increasingly are being used to estimate bedload transport. These estimates usually are derived by measuring changes in river channel form (Hicks and Gomez, 2016). However, terrigenous sediment also accumulates on Waimea Beach and the shoreline position has been monitored since 1926 (Fletcher et al., 2012). Additionally, the Kīkīaola Small Boat Harbor breakwater functioned as a dam, trapping sand on the beach after its construction in 1959 until its retention capacity was exceeded in the mid-1990s. Based on the ratio of volcanic to biogenic carbonate beach sediment, most (~90%) of this sand is derived from the Waimea River (Inman et al., 1963; Blay, 1999), with the remainder originating from the offshore fringing reef.

The position of the historical shoreline (interpreted low water mark) was determined at 20.1-m intervals along the 2.65-km-long beach and, based on the difference between the 1966 and 1992 shoreline positions, the annualized volumetric rate of change was 8188 m^3/y (Podoski, 2013). This equates to a mass input of 15.35 kt/y; 10% of which is assumed to be derived from the fringing reef. Some 63% of the flow in the estuary below the confluence with the Makaweli River is derived from the Waimea River and, assuming the rate of bedload transport also is a function of discharge, scaling the terrestrial sediment input to Waimea Beach by water discharge suggests the bedload discharge of the Waimea River is of the order of 8.64 kt/y (~23.7 t/d). Consistent with the uncertainty in the measured rate of shoreline accretion, an uncertainty of 50% is applied to this bedload discharge estimate.

Bedload often is assumed to constitute a relatively unimportant proportion (~10%) of the total physical load of energetic, high gradient rivers, although empirical evidence suggests that it may be considerably higher (Pratt-Sitaula et al., 2007). A high ratio of bedload to suspended load (>1:2) also has been observed in rivers draining basaltic terrain elsewhere that, because the main sediment sources lie outside the channel, scales with rainfall amount and increases as peak flow increases (Merten and Minella, 2014). Accepting that the suspended sediment load may be underestimated because the measurements were made during the warm phase of the PDO, a high ratio of bedload to suspended load can be sustained in the Waimea River because sand is readily mobilized by storm runoff on steep slopes and because a large amount of energy is available for particle transport.

The total riverine mass flux from the Waimea River basin (E_{fl}) is the sum of the bedload, suspended load, and solute fluxes. Expressed as a specific yield, to facilitate comparison with long-term estimates of the river's load (Gayer et al., 2008; Ferrier et al., 2013), $E_{fl} = 155 \pm 38$ t km^{-2} y^{-1} (Table 1).

5.1. Tempo of erosion in the Waimea River basin

The estimates of suspended and dissolved fluxes likely are underestimates because the warm PDO phase is associated with a decline in rainfall and streamflow, and because low flows that are significant for solute transport are diminished by agricultural diversions. For these reasons, neither measure is completely representative of the river's long-term, sediment transport regimes. However, North Pacific climate also varied throughout the Holocene, with a suppressed expression of a positive PDO and ENSO eventually being replaced by a more variable, El Niño-like and positive PDO-like climate pattern (Barron and Anderson, 2011). Thus, the modern-day relationship between precipitation and the ENSO and PDO climate oscillations also is unrepresentative of the average conditions that prevailed over the past 10 ky. Moreover, because the amount of chemical weathering that occurs in the near-surface environment on hillslopes is unknown, mass fluxes inferred from ^{3}He

Table 1
Riverine loads and fluxes from the Waimea River basin.

Applicable basin area[a] (km^2)	Timescale (y)	Load (t/d)	Load (kt/y)	Specific yield (t km^{-2} y^{-1})	Erosion rate[b] (mm y^{-1})	Comment
149.3	20	**30.0 ± 9.2**[c]	10.97 ± 3.36	74 ± 23	0.025 ± 0.008	Suspended
151.3	26		**8.64 ± 4.32**	57 ± 29	0.021 ± 0.010	Bedload
149.3	20	**10.1 ± 5.0**	3.70 ± 1.83	25 ± 12	0.008 ± 0.004	Dissolved
			23.31 ± 5.77	155 ± 38	0.052 ± 0.013	Total
130.0	10^4			168	**0.056**[d]	Gayer et al. (2008)
151.0	10^6			**140 ± 3**	0.047 ± 0.001	Ferrier et al. (2013)

[a] The initial estimates were derived from information obtained at different downstream locations.
[b] Density of basalt = 3 g/cm^3.
[c] Figures in bold type denote the initial estimates from which the other measures were derived.
[d] Average of spatially variable rates.

concentrations in olivine are open to a range of interpretations: from comparison with the sum of physical and chemical erosion rates to comparison with physical erosion rates alone (Ferrier et al., 2013). Nonetheless, bearing all the aforementioned caveats in mind, the modern-day erosion rate for the Waimea River basin closely approximates the ^{3}He-based erosion rate averaged over the past 10 ky (Table 1). That said, the measured (tenfold) variation in cosmogenic ^{3}He concentration implies that erosion rates across the Waimea River basin are far from uniform, and that the steepest hillslopes in the basin are eroding at rates that are more than an order of magnitude higher than the average rate (Gayer et al., 2008).

The modern-day erosion rate is faster than the volumetric-based rate averaged over the past ~4 My (Table 1). However, the erosional efficiency of the Waimea River and rivers throughout western Kaua'i positively correlates with mean annual rainfall (Ferrier et al., 2013). Based on a study of soil development on a remnant of the caldera's surface in the high-rainfall headwaters of the Kawaikōī watershed, because of climatic and topographic change mean annual precipitation over the last ~4 My is estimated as ~90% of its current value (Hotchkiss et al., 2000). Thus, in the absence of other (non-climatic) factors that strongly affect erosion rates and covary with mean annual rainfall (Ferrier et al., 2013), a slower My-scale erosion rate is consistent with a lower paleoprecipitation rate.

The Waimea River basin is the second fluvial system on Kaua'i where the riverine mass flux provides an estimate of modern-day erosion rates against which long-term erosion rate estimates can be compared. In the 47.9 km^2 Hanalei River basin the modern-day erosion rate (0.182 ± 0.043 mm y^{-1}) is three times faster than the erosion rate in the Waimea River basin, and also exceeds erosion rates averaged over both millennial and My timescales (Ferrier et al., 2013). Mean annual rainfall in the Waimea and Hanalei river basins is 1860 and 3866 mm/y respectively. The faster modern-day erosion rate in the Hanalei River basin may accrue from: (*i*) the covariation between erosion rates and rainfall; and (*ii*) the cumulative contribution made by sediment supplied by overland flow and shallow landsliding in the largely pristine, high-rainfall headwaters, and bank erosion and agricultural runoff lower down the basin (Takesue et al., 2009; Ferrier et al., 2013). Nonetheless, the modern-day erosion rates in both the Waimea and Hanalei river basins are lower than those found in river basins on other basaltic islands in the tropics (Rad et al., 2013). Why this is the case remains to be determined.

6. Why are the Waimea River's lower reaches aggrading?

River channels are formed, maintained, and altered by the water and sediment they convey. For a timescale of decades, a commonly used reference condition is an equilibrium stream profile that grades to a controlling base level. The slope of this 'graded stream' is adjusted to generate, with the available discharge and prevailing channel characteristics, the velocity required to transport the sediment load supplied from the drainage basin (Mackin, 1948). To maintain this condition, the potential a river has to perform geomorphic work (the product of water discharge and slope) tends to be in balance with the size and amount of sediment it transports (Lane, 1954), such that:

$$Q_s D \propto Q_w S \tag{1}$$

where Q_s is sediment load; D is sediment size; Q_w is water discharge; and S is river slope. Changing any of the terms on either side of the equation alters the balance, which can be redressed by a compensatory adjustment to one or more of the remaining variables. Even though the graded condition is a reference state, as opposed to a normative condition for channels, Eq. (1) has been used to make qualitative predictions about how rivers respond to changes in flow regime (Brandt, 2000; Phillips, 2013). The proportionality approximation also can assist with the process of river restoration (FISRWG, 1998; Booth and Fischenich, 2015). However, non-alluvial rivers, such as those flowing over bedrock, cannot adjust the size and quantity of the sediment they transport in this manner. Thus, Eq. (1) applies only to the Waimea River estuary and the adjoining upstream reach.

A more utilitarian form of Eq. (1) is:

$$q_s D^{3/2} \propto q_w^2 S^2 \tag{2}$$

where q_s and q_w are the sediment transport rate and water discharge per unit width (Henderson, 1966, p. 449). Solving Eq. (2) for S yields:

$$S \propto \frac{\sqrt{q_s} D^{3/4}}{q_w} \tag{3}$$

which can be used to compare conditions in the same reach, at different times (1, 2), if it is written as a ratio (Clark and Wilcock, 2000):

$$\frac{S_2}{S_1} = \left[\frac{q_{s2}}{q_{s1}}\right]^{1/2} \left[\frac{D_2}{D_1}\right]^{3/4} \frac{q_{w2}}{q_{w1}} \tag{4}$$

The slope in Eqs. (3) and (4) is that necessary to transport, with the available flow q_w, sediment supplied to a reach at rate q_s, and an increase in slope ($S_2/S_1 > 1$) is indicative of the tendency for sediment to accumulate within that reach under a changed regime (Clark and Wilcock, 2000).

Sediment production in the Waimea River basin depends on local topographic gradients, and most of the sediment in transport is derived from steep bedrock-dominated hillslopes where the size of the sand supplied to channels is presumed to be determined by the chemical weathering processes acting on them (Gayer et al., 2008; Sklar et al., 2017). Assuming q_s and D have not changed in recent decades, $q_{s2}/q_{s1} = 1$ and $D_2/D_1 = 1$. However, periods of heavy rainfall that cause direct runoff modulate streamflow, and any change in the frequency or magnitude of high streamflow events will affect the river's ability to transport sediment. After 1977, consistent with the warm PDO phase, there was a decline in the number of days when the mean discharge was above the threshold for sediment transport (~6 m^3 s^{-1}; Copeland, 1990), and in large (≥30 m^3 s^{-1}) winter floods. The magnitude of peak

annual flows in the Waimea River also declined by 22%. The size of the adjustment and the timeframe involved are indeterminate, but a flow reduction of this magnitude in the lower reaches of the Waimea River suggests that $q_{w2}/q_{w1} \approx 0.78$; thus $S_2/S_1 \approx 1.28$.

Most of the river's bed material load likely is transported by flows with a low (<1%) exceedance probability (Copeland, 1990). The agricultural diversions have a minimal impact on these flows (Fig. 3), but their magnitude-frequency characteristics are responsive to natural variations in climate. Although it is only a qualitative approximation, the implication is that a natural reduction in the magnitude of flows with low exceedance probabilities would have been sufficient to cause the lower reaches of the Waimea River to steepen through preferential deposition. The tendency for sand to accumulate under the changed hydrologic regime also would have been reinforced by the decline in the frequency and magnitude of very large floods (which reduced the river's ability to scour and remove this sediment) and vegetation growth on in-channel bars and benches.

The estuary also receives water and sediment from the Makaweli River, although peak flows in the two rivers coincide only ~50% of the time. There is a downward long-term, but not statistically significant, trend in peak annual flow in the Makaweli River. This basin's lower reaches are under pasture (Robinson Family Partners, 2016), but the three major tributaries all flow through deep, heavily vegetated canyons where rock avalanches that turn into muddy debris flows originate (Jones et al., 1984). Despite receiving more rainfall, the long-term erosion rate (121 ± 3 mm y^{-1}) is lower than in the Waimea River basin (Ferrier et al., 2013), but nothing is known about the Makaweli River's modern-day sediment transport regime.

The efficacy of fluvial processes declines to zero when a river reaches its controlling base level. In this case, the base level in question is the height of the Pacific Ocean surface. Differences in the flexural response of the lithosphere to volcanic loads across the Hawaiian island chain mean that each island experiences a unique rate of long-term sea level rise; for the southern shoreline of Kaua'i it is 17.5 ± 3.0 mm/decade (Fletcher et al., 2002). If rising sea level raises the lower end of a stream profile, there will be a concomitant adjustment to the vertical position of the river bed in closely adjoining reaches (Mackin, 1948; Lane, 1954; Leopold and Bull, 1979). From the one-dimensional, sediment continuity (Exner) equation:

$$dz/dt + dq_s/dx = 0 \quad (5)$$

where z is the bed elevation; t is time; and x is distance along the channel (Blum and Törnqvist, 2000); in a graded river $dz/dt = 0$ and $dq_s/dx = 0$. Thus, for a constant sediment supply, a rise in base level will be compensated for by a downstream decrease in the sediment transport rate and an increase in the rate of river bed aggradation. Aggradation also will occur if the controlling base level is displaced seaward (Lane, 1954), because the resultant reduction in slope diminishes a river's ability to transport sediment at a given discharge. This is effectively what happened after construction of the Kīkīaola Small Boat Harbor, as the shoreline adjacent to the river mouth accreted at an average rate of ~0.37 m/y between 1966 and 1992 (Podoski, 2013).

The implication is that rising sea level and seaward displacement of the shoreline have encouraged preferential deposition of an upstream tapering wedge of sediment in the estuary at the coast. Sea level is expected to continue to rise, but the position of shoreline adjacent to the river mouth appears to have stabilized since the harbor breakwater's retention capacity was exceeded in the mid-1990s.

Pōʻai Wai Ola hui's petition linked the "sediment problem" in the Waimea River estuary and the adjoining upstream reach to the diversion of water for off-stream uses. However, the large floods that tend to remove sand from the estuary were not diminished by the agricultural diversions during the plantation era; nor are they tempered by the much smaller amounts of water that currently are being diverted to support diversified agriculture on the Kekaha–Mānā Plain. Rather, it is a combination of extrinsic factors (a drier hydroclimate, sea level rise, and construction of the Kīkīaola Small Boat Harbor) that has promoted aggradation in recent decades. The cumulative impact of these external factors will neither be offset nor ameliorated by restoring any amount of flow to the Waimea River.

7. Conclusions and implications

Unusually complete hydrologic data permit the flow regimes of the Waimea River and the irrigation ditches that divert water from it to be characterized. The Waimea River also is the only place in Hawai'i (and one of a small, unique group of locations globally) where estimates of riverine loads derived from observations made over several decades can be compared to independent estimates of long-term erosion rates (Table 1). Collectively, the observational records span the last three and a half decades of the twentieth century. Flows between the threshold for sediment transport (~6.0 $m^3 s^{-1}$; Copeland, 1990) and the most effective flow (80.7 $m^3 s^{-1}$) transport ~60% of the Waimea River's suspended sediment load (Fig. 4B). Discharges of this magnitude recur annually and were unaffected by the plantation era agricultural diversions of 2.3 ± 0.7 $m^3 s^{-1}$ (Fig. 3).

Totaling the suspended sediment, bed, and solute flux estimates suggests that the modern-day riverine mass flux from the Waimea River basin is 155 ± 38 t $km^{-2} y^{-1}$. The low specific yield is consistent with the relatively low level of human interference with the natural environment and the first-order control weathering history exerts on erosion (cf. McLennan, 1993). Moreover, comparison of modern-day and independent ^{3}He-based estimates suggests that the basin-averaged erosion rate in the Waimea River basin has been in equilibrium with weathering conditions for the past 10 ky (Table 1).

Erosion rates across the Waimea River basin are far from uniform and the steepest, sparsely vegetated, bedrock-dominated hillslopes are eroding at rates that are more than an order of magnitude higher than the average rate (Gayer et al., 2008). These hillslope sources supply the sand and finer sized materials that the Waimea River transports to the ocean at a rate that is limited by the rate of sediment production. At very high discharges (recurrence interval > 2.5 years) the sediment transport regime of the Waimea River is supply-limited, and major floods tend to remove sand from the estuary (Copeland, 1990).

Climate change has caused periods of heavy rainfall to decline across the state (Kruk and Levinson, 2008; Seidel et al., 2008; Chu et al., 2010; Elison Timm et al., 2011; Garza et al., 2012), and the magnitude of peak flows in the pristine, undiverted headwaters of the Waimea River basin also has declined over the past 97 years (Fig. 2). If this trend endures in a warming, drying climate, incidences of heavy rainfall and the associated periods of high river discharge that scour the estuary will occur less frequently in the future.

The effect recent secular changes in climate have had on streamflow was presaged by the negative relation that exists between streamflow and the PDO (Oki, 2004). Lack of evidence for a long-term change in either the size or amount of sand in transport suggests that, beginning in the late 1970s, the naturally occurring, warm phase reduction in the magnitude of flows with low exceedance probabilities caused the lower reaches of the Waimea River to aggrade (Eq. (4)).

The controlling base level at the river mouth is the height of the Pacific Ocean surface, which is rising at a rate of 17.5 ± 3.0 mm/decade (Fletcher et al., 2002). After the Kīkīaola Small Boat Harbor was constructed and the adjacent shoreline accreted at an average rate of ~0.37 m/y (Podoski, 2013), sediment accumulated in the river mouth sand reservoir. This combination of rising sea level and seaward displacement of the shoreline created an upstream tapering wedge of sand that was preferentially deposited in the estuary at the coast (Eqs. 5 and 1).

After the harbor breakwater's retention capacity was exceeded, the position of shoreline stabilized; and this factor is unlikely to have an impact on future channel change. However, the fluvial system will

continue to respond to climate change brought about by global warming. That is, the tendency for sediment to be deposited in the river's lower reach will persist if periods of heavy rainfall and high river discharge continue to decline and sea level continues to rise.

Following the transition from plantation to diversified agriculture in the twenty-first century, the amount of water diverted into the ditch systems declined by 60%. Upstream agricultural diversions continue to impact low and intermediate flows in the Waimea River, and restoring additional flow to the river will further increase the magnitude of flows with a >10% exceedance probability. However, there is a threshold flow below which sediment does not move (~6 $m^3 s^{-1}$) and flows of this magnitude transport little or no sediment (Copeland, 1990). The flows that transport the most sediment are frequent events of moderate intensity that have a <10% exceedance probability. Flows of this magnitude are not modulated by agricultural diversions (Fig. 3), and neither they nor the very large flows that initiate scour will be appreciably amplified by the additional amounts of water that have been restored to the Waimea River.

The inescapable conclusion is that the anthropogenic stresses on Waimea River's lower reaches are not derived from the within-basin influence the agricultural diversions exert on the flow regime. Instead they accrue from extrinsic factors, the most profound of which will continue to influence the river's hydrologic and sediment transport regimes until global, offsetting, climate-ameliorating measures are implemented.

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